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INFLUX OF METEOR PARTICLES IN THE UPPER ATMOSPHERE OF THE EARTH AS DETERMINED FROM STRATOSPHERIC CORONAGRAPH OBSERVATIONS

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Abstract: This paper reports the observations of sky brightness near the sun at altitudes up to 25 km, using balloons. From the observed brightness the particle density is deduced as a function of altitude and divided into a component diffusing up from below and one settling from above. The deduced influx of particles in the size range of 0.1 to 2.8 microns radius is compared with rocket and other observations.

Резюме: В этой работе сообщаются данные наблюдений яркости неба вблизи Солнца на высотах выше 25 км, проведенные при помощи шаров-зондов. Из наблюдений яркости выводилась плотность частицы как функция высоты и разделялась на компоненты, диффундирующие сверху. Выведенный поток частиц радиусом 0·1–2·8 микрон сравнивался сданными наблюдения ракет и спутников.

1. Introduction

As part of a program to investigate the feasibility of coronal observations from balloons the High Altitude Observatory has flown an externally occulted coronagraph up to heights of 82 000 feet to measure the angular and the wavelength distribution of the daylight sky. As will be demonstrated in this paper, the sky radiance close to the sun is of interest not only to the instrument designer, who is concerned with the contrast between a target such as a planet or an artificial satellite and the sky at high altitudes, but also to the astronomer, who can make inferences about the sedimentation rate of meteoric material into the upper atmosphere. These inferences can serve as a valuable complement to the direct determinations of meteoric influx made from rockets and satellites.

2. Observations

The externally occulted coronagraph (fig. 1) used in this investigation was a modification of the instrument designed by Evans [1] and used

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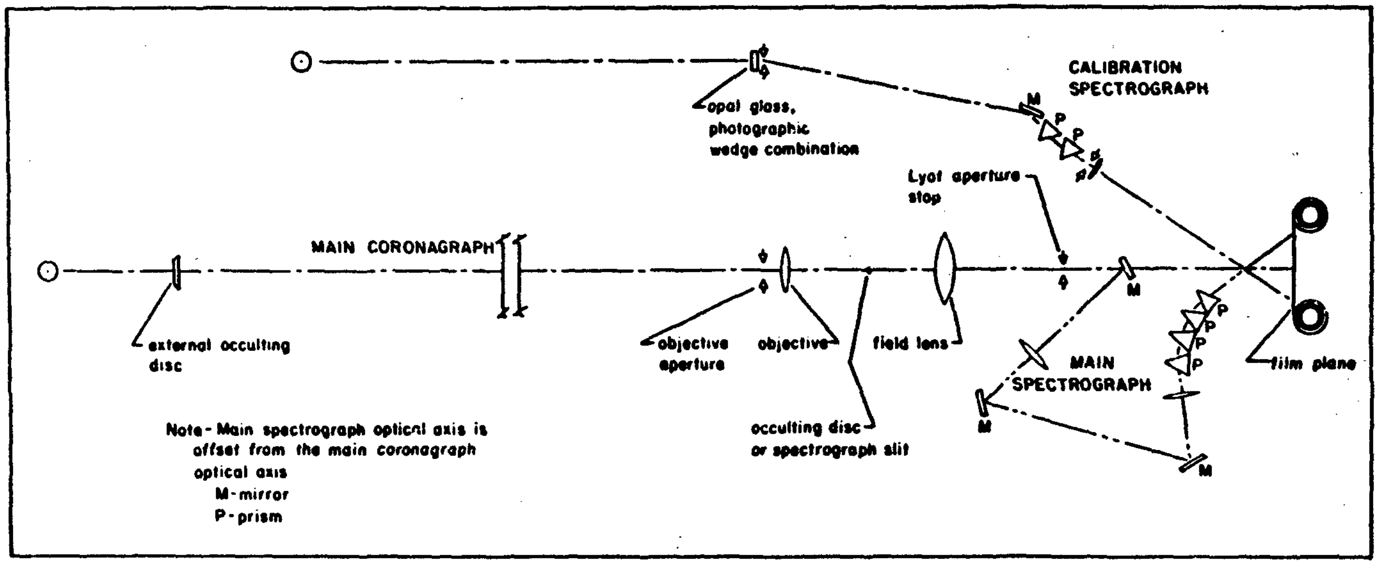

Fig. 1. Schematic diagram of the Flying Coronagraph. The external occulting disk, 150 cm from the objective lens, subtends an angle of 42 minutes of arc. Diffracted light from the external disk and the objective aperture is removed by the internal disk and the Lyot aperture, respectively. A small periscope (not shown) just below the external disk allows the objective to examine the skylight at scattering angles 9°6 to 57°8.

by one of us [2] for ground-based observations of the solar aureole. The instrument proved to introduce scattered light of radiance no higher than 10^{-8} of the mean radiance of the solar disk. This represented an improvement of approximately a factor of 200 over the ordinary coronagraph. The balloon-borne coronagraph was directed at the center of the solar disk to within an accuracy of $\pm$ 1 minute of arc by the guiding gimbals developed and used in Project Stratoscope I [3, 4, 5, 6]. The basic datum from the instrument consisted of sky radiances recorded in three ways: (1) photographically at scattering angles 9.°6, 20.°7, 31.°4, 40.°6, and 57.°8 from the center of the solar disk at $\lambda = 0.44\,\mu$ with electric vector vibrating perpendicular to the scattering plane, (2) photographically at scattering angles from 1.°67 to 2.°8 with a low dispersion spectrograph covering $\lambda\lambda\, 0.37\,\mu$ to $0.79\,\mu$, (3) photoelectrically at a scattering angle 10.°3 at $\lambda = 0.52\,\mu$. The spectrograph and the photoelectric photometer recorded total radiance. Standardization was provided by a sunlit, calibrated, opal glass and step-wedge combination which was photographed simultaneously with the spectrograph and the 9.°6 to 57.°8 fields. The photoelectric photometer was standardized frequently in flight with a calibrated opal illuminated by sunlight. An example of the variation of sky radiance with height at one scattering angle appears in fig. 2. The observations are compared with what would be expected, were the sky radiances measured in a pure molecular atmosphere with three different models of the radiation originating from below. Both the observations and the calculated values, taken from [7], include the effects of the variation of the solar zenith distance during the flight. Attention is called to the sharp approach of the observed sky radiance toward the calculated values above an altitude of 65 000 feet.

3. Interpretation

In order to interpret observations such as those shown in terms of the size distribution of aerosols above the balloon at any moment it is necessary to solve the equation governing the transfer of radiation through a turbid atmosphere. To this end, Sekera [8] and Deirmendjian [9, 10] have demonstrated the convenience of dividing the equation of transfer into two simultaneous equations – one describing the intensity due to scattering by the molecules and the other describing the intensity due to scattering by the aerosols. The first, governing the intensity in the pure molecular or Rayleigh atmosphere, has been solved by Chandrasekhar [11] and tabulated by Coulson, Dave, and Sekera [7]. By considering that, insofar as the overall transfer of radiation is concerned, the Mie scattering in the upper atmosphere

represents a small perturbation on the solution of the Rayleigh problem, we can use their tabulated solution. Subtraction of the radiance of the pure, molecular sky from that observed then yields the residual sky, which is

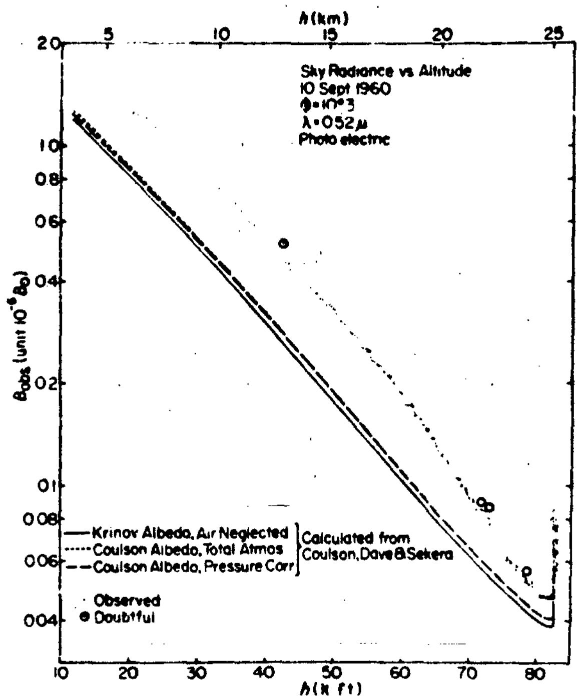

Fig. 2. Photoelectric observations of the sky radiance at a scattering angle of 10.°3 at $\lambda = 0.52\ \mu$. The "hook" at 82 000 feet was produced by the variation of solar zenith distance during the floating period of two hours. The observations are compared with three models of the radiation from a pure molecular sky.

attributed to scattering by the larger particles. By assuming that the irradiation of Mie particles by the rest of the atmosphere and by the ground is negligible and that $\mu \approx \mu_{\odot}$, we can write an approximate equation

$$\mu \frac{B_M{}^j}{B_{\odot}} = \omega_{\odot} \frac{\lambda^2}{8\pi^2} \int_{r_1}^{r_2} N(r, h)\, i^j (\alpha, \phi)\, \mathrm{d}r. \tag{1}$$

where

$\mu = \cos Z$,

Z – zenith distance of the point in the sky,

$B_M{}^j$ - radiance of the sky at height h, in the j^{th} polarization state, as due to the scattering by large particles,

$B_\odot$ = mean radiance of the solar disk observed at the height h,

$\omega_\odot$ = solid angle subtended by the sun,

h = height in the atmosphere.

$i^j(x, \phi)$ = Mie scattering intensity function in the j^{th} polarization state, for a spherical drop of size parameter x, of refractive index m, scattering through an angle ϕ (see [12] for a complete definition).

ϕ = scattering angle counted from the direction of forward scattering,

$x = 2\pi r/\lambda$ = size parameter of the droplet,

$N(r, h)$ = particle concentration in a column above height h with radii between r and $r+dr$.

The unit of N is cm^{-2} per micron dr. The usual convention is that $j=1$ refers to radiation with electric vector vibrating perpendicular to the scattering plane while $j=2$ refers to radiation with electric vector, normal to this plane. Calculation shows that the errors committed by using the

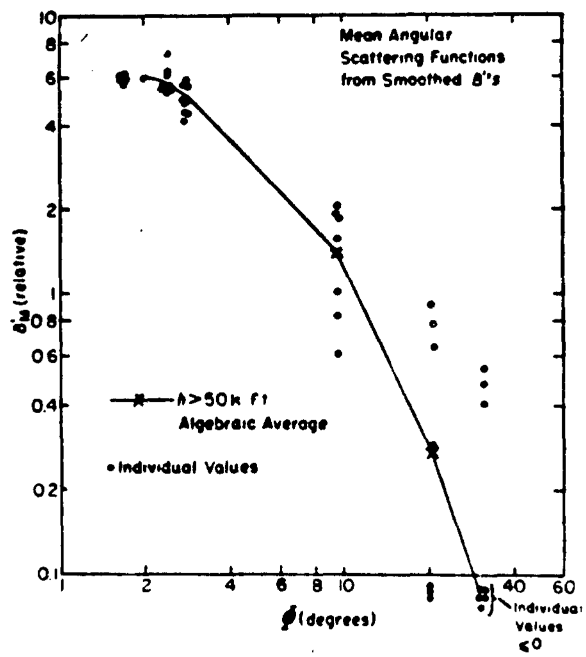

Fig. 3. The relative angular distribution of the light scattered by large particles for altitudes above 50 000 feet. All scattering diagrams have been normalized at $\phi = 2°$ a to relative intensity of 6. The algebraic mean of the measurements also appears.

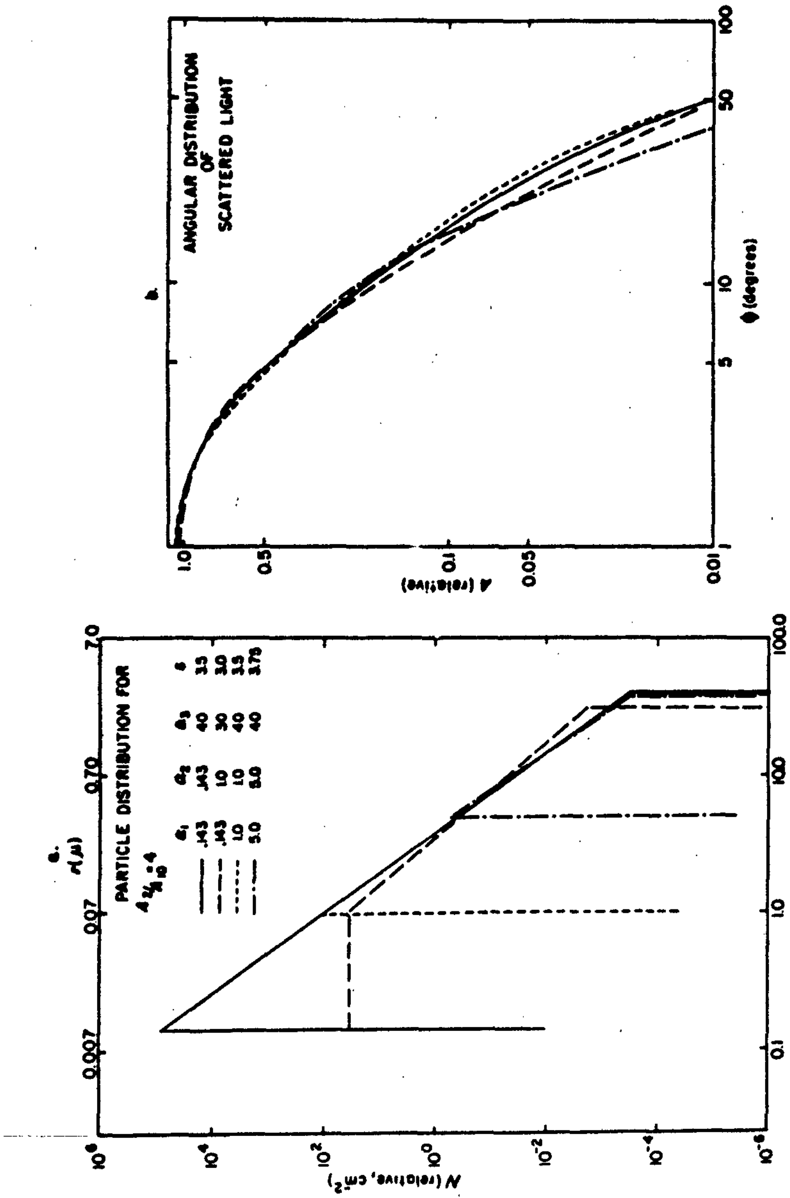

Fig. 4. Several particle size distributions (a) which produce nearly indistinguishable angular distributions (b) of scattered light. Here A is the relative scattered radiance at $\lambda = 0.44\ \mu$. Particle size distributions with δ much different from 3.5 in the $0.07\ \mu < r < 3.0\ \mu$ range produce scattering diagrams considerably different from those shown.

simplified equation in the stratosphere are of the order of 15 percent. Since most of the published Mie scattering coefficients are for $m = 1.33$, we have made our interpretation in terms of the scattering by spherical water drops. All particles are considered to have their radii between the limits r_1 and r_3.

The size distribution of the scattering aerosols is taken to have the form

$$N = N_0 \, (r/r_2)^{-\delta}, \tag{2}$$

where

$$\delta = 0, \quad r/r_2 < 1,$$
$$\delta = \delta, \quad r/r_2 > 1.$$

Junge [13] and his collaborators [14, 15] have found by direct sampling, and Volz [16], Newkirk [2], Deirmendjian [9, 10], and Volz and Bullrich [17] have inferred from optical measurements that such a form adequately describes the size distribution of atmospheric aerosols. In principle, the

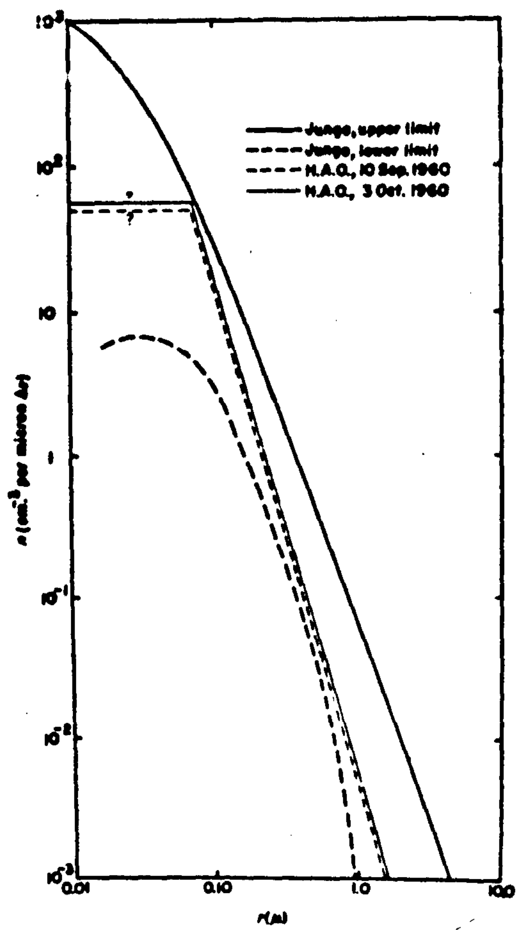

Fig. 5. A comparison between the particle size distribution at 66 000 feet determined optically on 10 September 1960 and on 3 October 1960 and the limits of particle concentration set by Junge and his collaborators by direct sampling techniques. The optical determination is ambiguous for $r < 0.1\ \mu$.

angular distribution of the residual sky, such as appears in fig. 3, should determine the parameters r_1, r_2, r_3, and δ of the size distribution. In fact, however, a certain ambiguity is inherent in the problem. This is demonstrated in fig. 4 in which the angular variation of scattered light from aerosols with different size distributions is demonstrated. A comparison between the observed angular distribution of scattered light and that calculated shows that the models displayed could hardly be distinguished on the basis of the observations. Clearly, the observations allow nothing to be said concerning the size distribution below a radius of approximately 0.1 μ. Between $r = 0.1\ \mu$ and 3.0 μ, however, the size distribution is fairly well determined, and the parameter δ is established as $3 < \delta < 4$. Examination of the wavelength distribution of the residual sky does little to remove the ambiguity concerning the smaller particles. In spite of these difficulties, the optical data yield a size distribution for particles in the upper atmosphere which is quite consistent with that obtained by direct sampling (fig. 5).

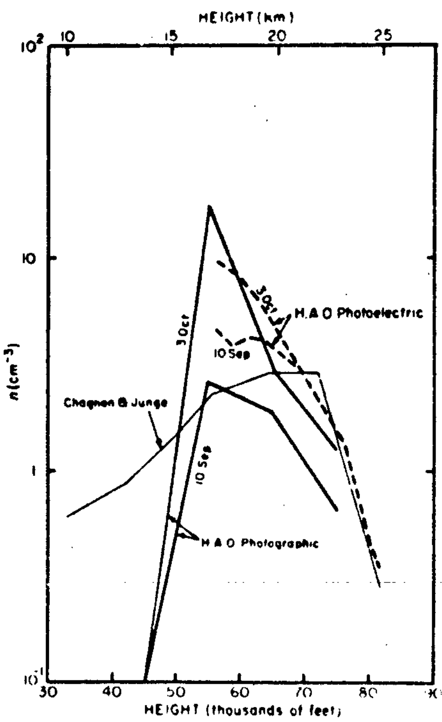

Fig. 6. Observed variation of particle concentration with height for $r = 0.15\ \mu$ determined optically on 10 September 1960 and 3 October 1960 and by Junge and collaborators by direct sampling.

By differentiation of the size distributions with respect to height, one can determine the particle concentration at any particular altitude. The result of such an analysis appears in fig. 6, in which the variation with height of particles of radius 0.15 μ is shown. This particular radius for the particle was chosen in order to allow a comparison with the direct sampling results. Quite clearly, our observations confirm the conclusion of Junge and his collaborators that an aerosol layer at a height of approximately 65 000 feet exists. The form of such a layer suggests that the majority of particles in it are of terrestrial origin and that the maximum is due either to the concentration of particles by advection or by *in situ* formation.

4. The net vertical flux of particles

The sharp drop of particle concentration with height above the maximum of the aerosol layer can be further interpreted under the assumption that a steady state prevails and only two mechanisms -- diffusion and sedimentation – govern the flow of particles. Junge and his collaborators [14] write the vertical flux of particles γ as

$$\gamma = - n_s w_s - n \frac{d\nu_s}{dz} D, \tag{3}$$

where

n_s = concentration of particles of radius r,
w_s = fall speed of the particles of radius r,
n = concentration of air molecules (number per cm³),
ν_s = concentration of particles relative to the concentration of air molecules,
D = diffusion parameter.

The fall speed is that given by the Stokes–Cunningham relation [18] while the diffusion parameter contains the effects of both eddy and molecular diffusion. Within the layers of the atmosphere of concern in our problem, eddy diffusion is dominant [19]. The solution of eq. (3) in an atmosphere of scale height H yields

$$n_s = n_{s0}\, e^{-z/H - \int_0^z (w/D)\, dz'} - \frac{\gamma}{w} \left\{ 1 - e^{- \int_0^z (w/D)\, dz'} \right\}, \tag{4}$$

where the concentration is n_{s0} at $z = 0$. The effect of increasingly large values of the diffusion constant D on the variation of particle concentration with height appears in fig. 7, in which the diffusion is considered to occur from a height 66 000 feet. By interpreting the observed drop in particle con-

centration with height above the maximum of the aerosol layer as due to the mechanism of sedimentation-diffusion, we can determine that the diffusion constant D from 66 000 to 80 000 feet lies in the range 2000 cm²/sec $< D <$ 10 000 cm²/sec with the best fit for $D \approx 5000$ cm²/sec. This value is in rough agreement with the $D \approx 10^3$ estimated by Lettau [19] for temperate latitudes from models of the representative zonal circulation and the lapse rate. The fact that the slow decrease in concentration of the curve m (fig. 7) is not observed indicates that the majority of the particles in this height range are not of meteoric origin.

However, the additional assumption that D is constant above 80 000 feet allows the calculation of the total concentration (in the cm² column) overhead of particles with radius r which have been carried aloft by diffusion. (In fact, the usual profile of temperature *vs* height above 80 000 feet would

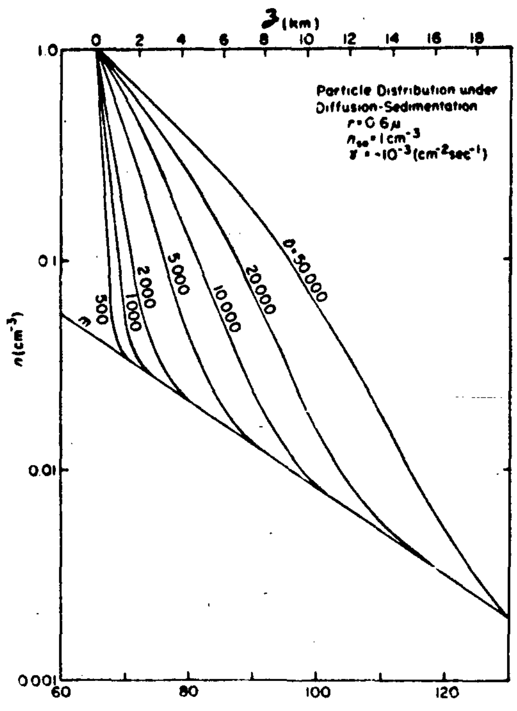

Fig. 7. Variation of relative particle concentration with height above the maximum of the aerosol layer (taken at 66 000 feet) to be expected if sedimentation-diffusion equilibrium applies. A constant influx of meteoric material is assumed, and the concentrations are calculated for several values of the diffusion constant D. If the only source of particles were the sedimentation of meteoric dust with influx rate γ, the concentrations m would be observed.

lead to the expectation that D should decrease slightly with height.) This calculation yields column concentrations for $0.1\,\mu < r < 3\,\mu$ which are below those required by the scattering observations. In other words, the air above 80 000 feet contains more particles in this size range than can be explained by diffusion from below. The origin of the excess particles is to be found in the sedimentation of meteoric material downward. Thus, for a particular value of D, the observations lead to a value of the meteoric influx γ. In fig. 8 appears a comparison between the meteoric influx derived in this manner for $0.1\,\mu < r < 3.0\,\mu$ and as determined by various rocket experiments. The balloon measurements agree with the more conventionally determined meteor fluxes but are somewhat divergent from the "Venus Fly Trap" [20]

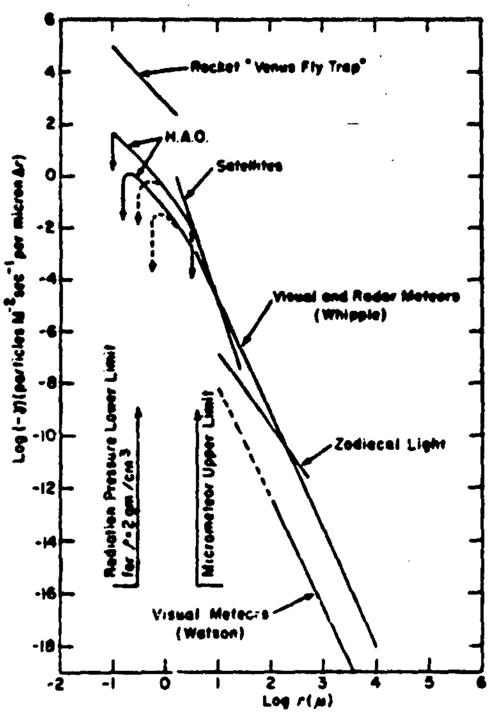

Fig. 8. Comparison of meteoric influx rates inferred in this paper with influx rates inferred from the rocket "Venus Fly Trap" [20], satellites [21], visual and radar meteors [22], zodiacal light [23] and visual meteors [24]. The higher HAO curves refer to the results of 10 September 1960; the lower curves, to 3 October 1960. For $D = 2000$ cm^2/sec the solid HAO curves apply, while for $D = 10\,000$ cm^2/sec the dashed curves apply. The lower limit for particles to remain in the solar system against radiation pressure and the upper limit for particles to survive the passage through the atmosphere [25] are indicated.

results. The cause of this difference is unknown. Our analysis also seems to confirm the calculation of Whipple [25], that only particles with radii greater than about 4 μ are destroyed in passing through the atmosphere.

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